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May 1977

A PROPAGATING LARGE-AMPLITUDE DISTURBANCE IN A STRATIFIED OCEAN WITH A TEMPERATURE INVERSION

By: F. J. DYSON

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This paper presents a hypothetical mode of propagation of a large-amplitude disturbance involving turbulent mixing in a temperature-inverted layer. The motions are described empirically, calculating a velocity of propagation consistent with conservation of mass and momentum for a travelling disturbance of the prescribed pattern. This results in propagation velocities of the order of one knot under realistic oceanic conditions.

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ABSTRACT

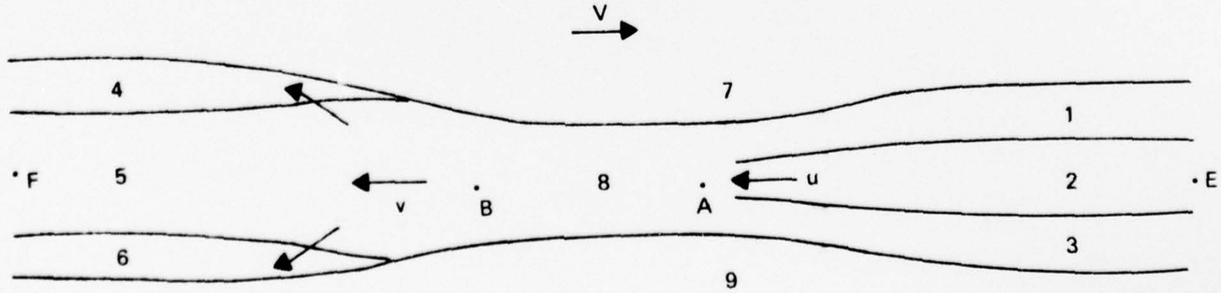
This paper presents a hypothetical mode of propagation of a large-amplitude disturbance involving turbulent mixing in a temperature-inverted layer. The motions are described empirically, calculating a velocity of propagation consistent with conservation of mass and momentum for a traveling disturbance of the prescribed pattern. This results in propagation velocities of the order of one knot under realistic oceanic conditions.

A PROPAGATING LARGE-AMPLITUDE DISTURBANCE IN A STRATIFIED OCEAN WITH A TEMPERATURE INVERSION

It happens quite frequently that a temperature inversion persists for long periods in a stably stratified ocean, for example when a layer of warm salty water lies between two layers of cool less-salty water of roughly equal density. Such a situation is stable against small perturbations. But it may be unstable against large disturbances, because a turbulent mixing of the layers can result in segregation of cool salty water and warm less-salty water, as demonstrated in a tank experiment by J. S. Turner and C. F. Chen, "Two-dimensional effects in double-diffusive convection," J. Fluid Mech. 63, 577 (1974), (see especially page 588 and Plate 10). There is free energy available to drive an instability, if a suitable mechanism exists for releasing the energy.

The purpose of this note is to describe a hypothetical mode of propagation of a large-amplitude disturbance involving turbulent mixing in a temperature-inverted layer. I do not attempt to prove that this mode of propagation follows from the exact laws of hydrodynamics. I only describe the motions in an empirical fashion, and calculate a velocity of propagation that is consistent with conservation of mass and momentum for a traveling disturbance of the prescribed pattern. It turns out that propagation velocities of the order of 1 knot are to be expected under realistic oceanic conditions.

The figure gives a schematic representation of the flow pattern in the propagating disturbance. The pattern is supposed to propagate from left to right with velocity v, while the water moves from right to left through the throat in region 8. The initial state of the water is seen at the right, with the three layers 1, 2, 3 having equal density ρ, layer 2 being warm and salty while 1 and 3 are cool and less salty. The upper layer 7 with density $\rho(1 - \beta)$ and the lower layer 9 with density $\rho(1 + \beta)$ are assumed not to mix with the active layers 1, 2, 3. The final state of the water is seen at the left, after layers 1, 2, 3 have become turbulently mixed and the differential diffusion of heat and salt has resulted in the segregation of the cold salty layer 6 and the warm less-salty layer 4. I assume that layers 4, 5, 6 have densities $\rho(1 - \alpha)$, ρ and $\rho(1 + \alpha)$, respectively. Let D be the combined thickness of layers 4, 5, 6, which is equal to the combined thickness of layers 1, 2, 3. The thicknesses of layers 4 and 6 are assumed to be each equal to xD, where x is a fraction depending on the details of the double-diffusive process.



There is now a hydrostatic pressure difference

$$\Delta P = g\rho\alpha x D \qquad (1)$$

between the points E,F at the same depth in layers 2 and 5. It is this pressure difference that drives the movement of water from right to left. The movement takes place in three phases. First, there is a laminar flow of the three equal-density layers 1, 2, 3 up to the point labeled A in the Figure. I assume for this laminar flow a roughly parabolic profile, so that the velocity u at the central point A is larger than the average velocity v of the flow. The second phase of the flow is the growth of a Helmholtz instability followed by complete turbulent mixing of layers 1, 2, 3 in the throat region 8. The third phase is the movement of the homogenized water past point B with velocity v, and its coming to rest in the segregated layers 4, 5, 6. We have

$$v = fu \quad , \qquad (2)$$

where f is a fraction equal to 2/3 for a parabolic velocity profile.

Let the thickness of the water in region 8 be

$$D' = (1 - 2y)D \quad . \qquad (3)$$

The velocity u is related to the pressure difference between A and E by Bernoulli's equation

$$\tfrac{1}{2}\rho u^2 = g\rho\beta yD \quad , \tag{4}$$

and similarly v is related to the pressure difference between B and F by

$$\tfrac{1}{2}\rho v^2 = g\rho(\beta y - \alpha x)D \quad . \tag{5}$$

Finally, the propagation velocity V is determined by the mass-conservation equation

$$(V + u)D' = VD \tag{6}$$

We have five equations (2) - (6) for the five unknowns u, v, V, y, D', the parameters f, D, ρ, α, β, x being assumed given by the initial conditions of the problem.

From Eq. (2), (4), (5) we obtain the solution

$$y = \alpha x/(\beta(1 - f^2)) \quad , \tag{7}$$

$$u = (1 - f^2)^{-\frac{1}{2}} (2g\alpha xD)^{\frac{1}{2}} \quad , \tag{8}$$

and then V is obtained from Eq. (3) and (6),

$$V = \left[\frac{\beta(1-f^2)}{2\alpha x} - 1\right]\left[\frac{2f^2 g\alpha x D}{1-f^2}\right]^{\frac{1}{2}} \quad . \tag{9}$$

The quantitative details of Eq. (9) are of course not be taken seriously. The important fact is that the order of magnitude of the propagation velocity is determined by

$$V \sim (g\alpha D)^{\frac{1}{2}} \quad , \tag{10}$$

as could have been foreseen from a simple dimensional argument. The fractional density difference α resulting from the double-diffusive separation may be expected to be of the order of magnitude

$$\alpha = 10^{-4}\,\Delta T \quad , \tag{11}$$

where ΔT is the magnitude of the initial temperature inversion in degrees Celsius.

As an illustration, I take in Eq. (9) the numerical values

$$f = \frac{2}{3} \quad , \quad x = \frac{1}{10} \quad , \quad \beta = \alpha \quad , \tag{12}$$

and obtain the result

$$V = (5\alpha D)^{\frac{1}{2}} \quad , \tag{13}$$

with D in meters and V in meters per second. Then with $\Delta T = 5^{\circ}$ and $D = 100$ meters, we have finally

$$V = 0.5 \text{ meter/sec} = 1 \text{ knot} \quad . \tag{14}$$

This is only a rough estimate, but it would seem to be difficult to obtain a propagation velocity as large as 3 knots without stretching the parameters to rather extreme values.

On pages 271 and 272 of J. S. Turner's book "Buoyancy Effects in Fluids" (Cambridge University Press, 1973), he shows examples of actual temperature and salinity inversions observed in the ocean. The inverted layers have an overall thickness of several hundred meters, subdivided into many sub-layers by double-diffusive convection. In this note I have ignored the additional complications introduced by the sub-layers and treated the whole inverted layer as homogeneous. In a correct theory of traveling disturbances in an inverted layer, the sub-layering phenomena should certainly be taken into account.

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